8-1996

Influences of Model Parameterization Schemes on the Response of Rainfall to Soil Moisture in the Central United States

Zaitao Pan
Iowa State University

Eugene S. Takle
Iowa State University, gstakle@iastate.edu

Moti Segal
Iowa State University

Richard W. Turner
Iowa State University

Influences of Model Parameterization Schemes on the Response of Rainfall to Soil Moisture in the Central United States

ZAITAO PAN

Department of Geological and Atmospheric Sciences, Iowa State University, Ames, Iowa

EUGENE TAKLE

Department of Geological and Atmospheric Sciences and Department of Agronomy, Iowa State University, Ames, Iowa

MOTI SEGAL

Department of Agronomy, Iowa State University, Ames, Iowa

RICHARD TURNER

Department of Geological and Atmospheric Sciences, Iowa State University, Ames, Iowa

(Manuscript received 31 July 1995, in final form 24 January 1996)

ABSTRACT

The sensitivities of soil moisture impacts on summer rainfall in the central United States to different commonly used cumulus parameterization and surface flux schemes are examined using the PSU–NCAR MM5 under different atmospheric and soil moisture conditions. The cumulus convection schemes used are the Kuo and Grell parameterization schemes, while the surface-moisture flux schemes used are the aerodynamic formulation and the Simple Biosphere (SiB) Model. Results show that a transient increase in soil moisture enhanced total rainfall over the simulation domain. The increase in soil moisture enhanced local rainfall when the lower atmosphere was thermally unstable and relatively dry, but it decreased the rainfall when the atmosphere was humid and lacked sufficient thermal forcing to initiate deep convection. Soil moisture impacts were noticeably stronger for the Kuo scheme, which simulated lighter peak rainfall, than those for the Grell scheme, which simulated heavier peak rainfall. The greater sensitivity to soil moisture exhibited by the Kuo scheme than either the Grell or explicit schemes implies that it exaggerated the role of soil moisture. This difference was related to how each scheme partitioned rainfall between convective and stable forms, and possibly to each scheme's closure assumptions. Adding details to the surface-moisture flux schemes had a secondary influence on soil moisture impacts on rainfall within a 24-h period.

1. Introduction

Surface moisture fluxes over land are an important forcing for the atmosphere over continents on a wide range of spatial and temporal scales. Over the last three decades, climate modelers have examined and quantified soil moisture effects on precipitation climatology. In a pioneering general circulation model (GCM) simulation by Manabe (1969), the importance of surface hydrological processes to the atmospheric circulation was illustrated. Subsequent GCM simulations (e.g., Mintz 1984; Rowntree and Bolton 1983) showed that soil moisture contributed positively to global rainfall. With increasing interest in regional climate modeling (Giorgi et al. 1993), recent soil moisture impact modeling studies have tended to concentrate on regional scales for shorter time periods, (e.g., a month or a season). For example, Atlas et al. (1993) simulated the effects of soil moisture on rainfall over the Great Plains during the North American drought of 1988.

Monthly total rainfall on regional or local scales is mainly determined by a few heavy rainfall events, creating inhomogeneous regional soil-moisture patterns that may persist for a few weeks or longer. These isolated heavy-rain events may have different climatic impacts than spatially or temporally averaged patterns. Recent studies (e.g., Benjamin and Carlson 1986; Chang and Wetzel 1991; Lakhtakia and Warner 1987; Lanicci et al. 1987; Yan and Anthes 1988) evaluated the combined impact of inhomogeneity in surface wetness and dynamic processes on deep convection on short temporal scales. A preliminary evaluation of tran-

Corresponding author address: Dr. Zaitao Pan, 3010 Agronomy Hall, Dept. of Geological and Atmospheric Sciences, Iowa State University, Ames, IA 50011.
E-mail: panz@iastate.edu

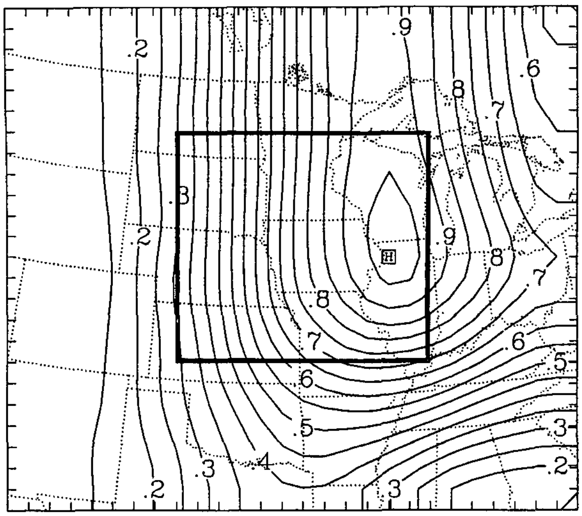

FIG. 1. Simulation domains and estimated soil-moisture availability *m* field for 9 July 1993. The nested domain is indicated by the inner frame.

sient soil moisture effects on rainfall was carried out by Pan et al. (1995) using coarse model resolutions and relatively simple physical parameterizations of cumulus convection and surface fluxes. Segal et al. (1995) also scaled the dependency of local deep convection on the Bowen ratio over uniform surfaces.

The credibility of modeled soil moisture impacts on rainfall depends on the correctness of simulated rainfall. Unfortunately, rainfall is one of the most difficult variables to simulate due to contributions from subgrid-scale (i.e., unresolvable) convection and cloud microphysical processes (Molinari and Dudek 1992). Because of the uncertainties involved in parameterizing these processes, simulated precipitation amounts and patterns can vary widely from scheme to scheme (Zhang et al. 1988; Wang and Seaman 1994). For example, since the Kuo cumulus convection scheme usually underestimates peak rainfall, it will be useful to determine whether the soil-moisture impacts on subsequent rainfall also are underestimated if the Kuo scheme is used. Likewise, the surface flux schemes for evapotranspiration (ET), which regulates the surface moisture flow to the atmosphere, should influence soil moisture impact evaluation. As an example, the most commonly used aerodynamic formulation tends to overestimate evaporation (e.g., Sato et al. 1989). Therefore, it is important to determine to what extent this excess moisture influences future rainfall.

The intercomparison of models as a means of promoting progress in atmospheric modeling has been encouraged by the Working Group on Numerical Experimentation of the World Climate Research Program (WGNE 1992). The present paper evaluates the sensitivities of soil moisture impacts on rainfall by intercomparing different cumulus parameterization schemes and different surface flux schemes over three separate 24-h periods. A wide range of soil moisture regimes is examined to resolve the nonlinear relationship of rainfall to soil moisture.

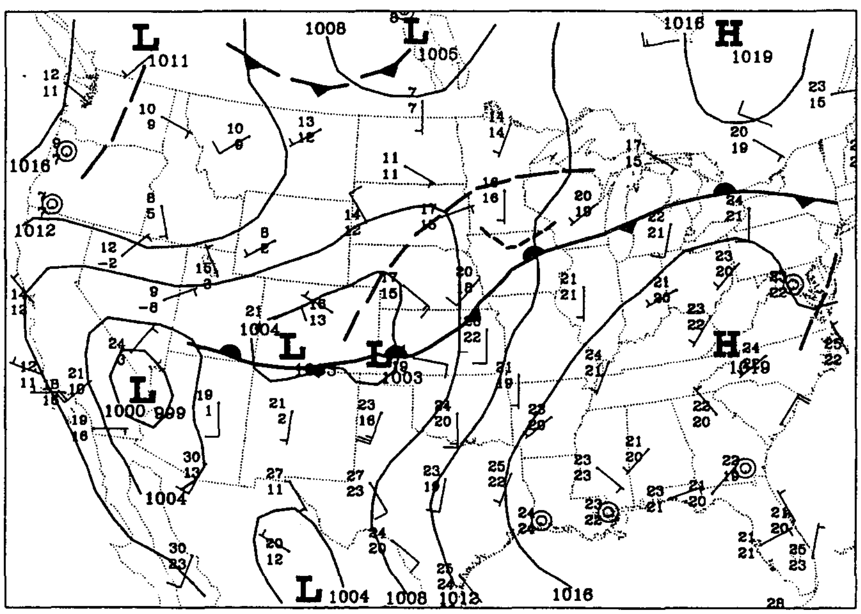

FIG. 2. Surface synoptic chart at 1200 UTC 8 July 1993 (adapted from the *Daily Weather Maps* issued by the Climate Analysis Center of the NMC). The long-dashed lines are troughs and the short-dashed line is an outflow boundary.

Section 2 briefly describes model schemes used in this study. In section 3, sensitivities of soil moisture impacts on rainfall to cumulus parameterization schemes and to surface flux schemes are examined for the base case (9 July 1993). Soil moisture impacts on evapotranspiration are evaluated in section 4 for different surface flux schemes. In section 5, the sensitivity of soil moisture impacts on rainfall is examined for two additional synoptic cases during the 1993 flood.

2. The mesoscale model and parameterization schemes

a. Mesoscale model

The fifth-generation PSU–NCAR mesoscale model (MM5) (Grell et al. 1993) was used for the present study. The main advantages of MM5 over its previous version MM4, which has been used widely, are its nonhydrostatic option and more detailed model physics formulations incorporated in the cumulus convection, radiative transfer, and cloud schemes.

The nonhydrostatic option and explicit warm cloud physics (Hsie and Anthes 1984) were employed in this study. This explicit moisture scheme operates together with a cumulus parameterization to form a so-called hybrid scheme for the calculation of precipitation. A one-level nested grid was used in which the outer coarse domain had 28 (east–west) × 25 (north–south) grid points with 90-km resolution and the inner nested domain had 37 (east–west) × 34 (north–south) grid points with 30-km resolution. The outer domain covered most of the central and eastern United States, while the nested domain was centered over the heavy rainfall area for the case (as indicated in Fig. 1). Twenty-three layers in the vertical were used for both domains with about 5-hPa resolution near the ground. Because this study emphasizes the transient response of rainfall to soil-moisture change, the period of all simulations was 24 h starting at 1200 UTC of the previous day.

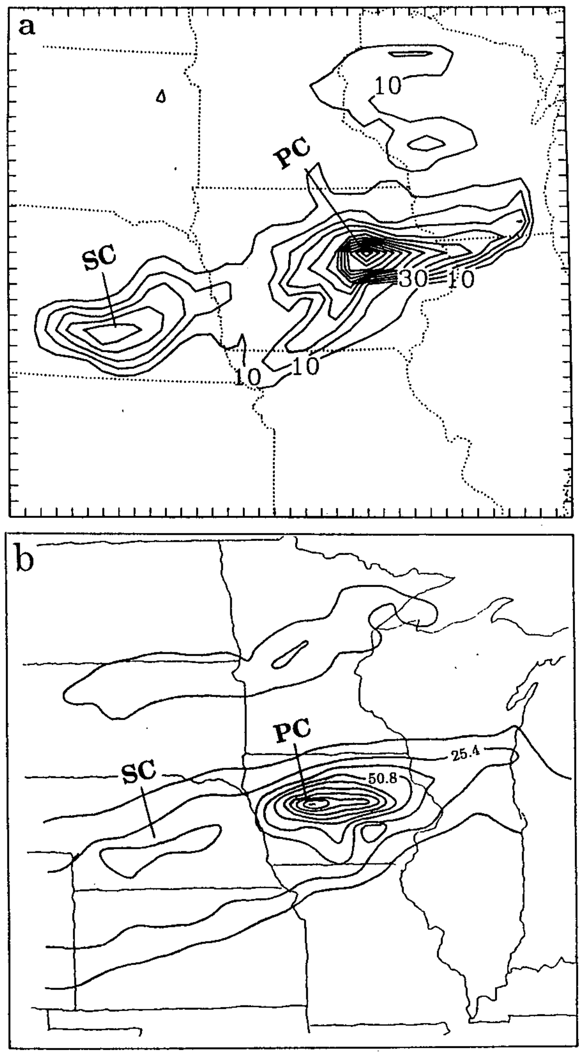

FIG. 3. The 24-h accumulated rainfall fields (mm) at 1200 UTC 9 July 1993: (a) simulated (contour interval 10 mm); (b) observed [adapted from Richards and Hudlow 1994; contour interval 12.7 mm (0.5 in.)]. PC and SC denote the primary and secondary rainfall centers, respectively.

b. Cumulus parameterization schemes

The two cumulus parameterization schemes used in this study were the Kuo scheme, denoted KS (Kuo 1974; Anthes 1977), a widely used cumulus parameterization, and the Grell scheme, denoted GS (Grell 1993), modified from the Arakawa–Schubert scheme (Arakawa and Schubert 1974) that has a more sound physical base (Raymond and Emanuel 1993). In KS, which does not explicitly account for downdraft circulation, convective precipitation P is expressed as

$$P = M_t(1 - b), \tag{1}$$

where M_t is the vertically integrated moisture convergence in a grid column including ET and b is a moistening factor that is a function of relative humidity (Anthes 1977). The characteristics of KS have been widely tested (e.g., Kuo and Anthes 1984; Wang and Seaman 1994). In GS, which explicitly incorporates a single downdraft plume of clouds, convective precipitation is expressed as

$$P = I_1 m_b(1 - \beta), \tag{2}$$

where I_1 is the total condensate normalized by mass flux in the updraft, m_b is the mass flux at cloud base, and β is the fraction of condensate detrained out of clouds and reevaporated, which is determined by vertical wind shear (Fritsch and Chappell 1980). The performance of GS has been evaluated by Giorgi (1991) and Wang and Seaman (1994). The differences in precipitation physics formulations between these two schemes are quite substantial

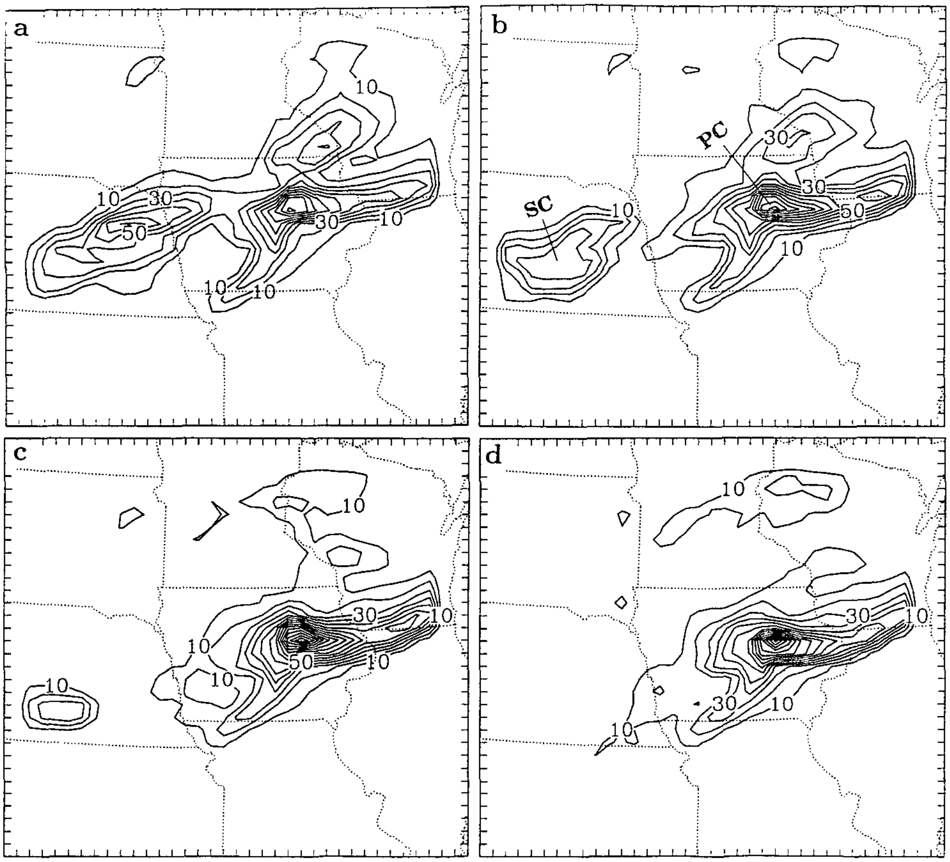

FIG. 4. Simulated 24-h accumulated rainfall (mm) at 1200 UTC 9 July 1993 using GS as the cumulus convection scheme and AD as surface flux scheme for different m values: (a) $m = 0.1$, (b) $m = 0.2$, (c) $m = 0.4$, and (d) $m = 0.6$. PC and SC denote the primary and secondary rainfall centers, respectively.

even though the final expressions for precipitation in Eqs. (1) and (2) seem to be similar.

c. Surface moisture flux schemes

The conventional aerodynamic (AD) formulation and the more detailed Simple Biosphere (SiB) Model (Sellers et al. 1992; Collatz et al. 1990) were used to intercompare the dependence of rainfall on soil moisture under different surface flux parameterization schemes.

The standard version of MM5 computes ET (surface moisture flux) using an AD formulation (Blackadar 1976; Zhang and Anthes 1982). The evapotranspiration is given by

$$\mathrm{ET} = m\rho_a C_q V_a (q_{gs} - q_a), \qquad (3)$$

where q_{gs} is the saturation mixing ratio with respect to surface soil temperature (T_g), ρ_a, V_a, and q_a are the air density, wind speed, and water vapor mixing ratio at the lowest model level, respectively, and C_q is the surface exchange coefficient for moisture. The soil moisture availability m is the ratio of actual ET to potential ET. A value of $m = 0$ denotes completely dry top soil and a value of $m = 1$ represents saturated top soil. (Soil moisture hereafter refers to m, unless otherwise indicated.) ET is a nonlinear function of m (most noticeably for low values of m) because of the exponential dependence of q_{gs} on T_g.

SiB2 (a newer version of SiB) was incorporated into MM5 for this study. In SiB2, ET consists of transpiration from the canopy (E_c) and evapotranspiration from the ground cover (E_g). Soil moisture (representing the average soil moisture in SiB2) affects ET by controlling stomata and soil resistances. (An outline of the ET computation in SiB2 is given in the appendix.)

3. Sensitivity of soil-moisture impacts on rainfall to parameterization schemes—9 July 1993 case

In this section, the distribution of rainfall simulated by using different schemes will be examined under various soil-moisture conditions.

a. Synoptic situation

The midwestern United States experienced its worst flood on record in the summer of 1993 (Gerald and

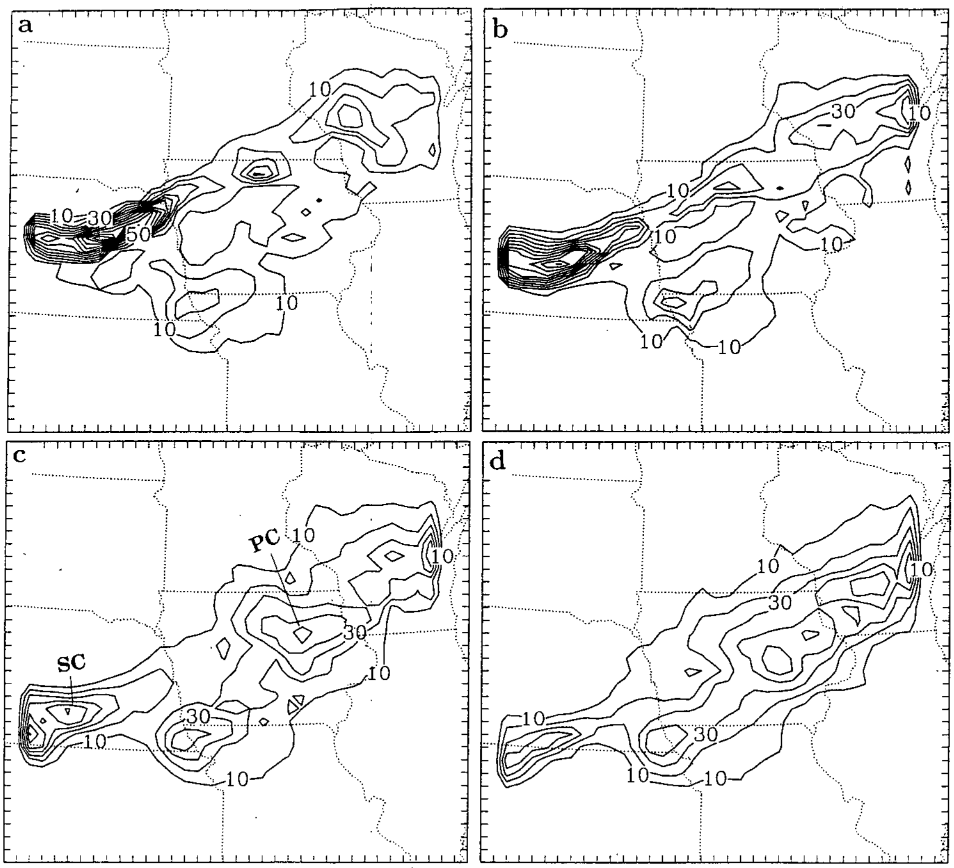

FIG. 5. Same as Fig. 4 except for using KS as the cumulus convection scheme.

Janowiak 1995). This long-lasting and widespread flood (termed the Great Flood) resulted from a series of heavy rainfall events during June and July (Rodenhuis et al. 1994; Mo et al. 1995). Three days (17 June, 5 July, and 9 July) with peak daily rainfall exceeding 100 mm were simulated.

The 9 July case was the heaviest rainfall event, and thus it serves as the focal (base) case for detailed analysis. A peak rainfall of 150 mm was observed in central Iowa over the 24-h period ending at 1200 UTC 9 July. An unseasonably high amplitude upper-air pattern characterized by a strong ridge over the Gulf of Alaska and a deep trough over the western United States existed during this period. East of the trough, that is, over the central United States, there was a strong baroclinic zone and an upper-level jet with a strong southerly component.

At the surface a quasi-stationary front was positioned through Iowa, extending from western Kansas to central Michigan. A high pressure system anchored over the southeastern United States in conjunction with an area of low pressure over western Kansas and Colorado produced broad southerly surface flow that advected warm moist air into the central United States (Fig. 2). A low-level jet (LLJ) was also present above the region of southerly surface winds and this enhanced the moisture transport into the region and it provided a source of convective instability. The presence of the jet stream aloft, the LLJ, and outflow boundary remnants from storms of the previous night provided extremely favorable conditions for the heavy rainfall event that occurred (Augustine and Caracena 1994).

b. Sensitivity of simulated rainfall to cumulus parameterization schemes

1) RAINFALL FIELDS

Soil moisture availability m for the base case was estimated from the weekly crop moisture index and Palmer drought index (issued by the NWS) for the second week of July 1993 (Fig. 1). Persistent flooding produced an extremely wet surface with maximum m values above 0.9 over much of the upper Mississippi River basin. On the other hand, low m values existed in the southeast corner of the domain and in the Rocky Mountain region.

Figure 3a shows the simulated 24-h accumulated rainfall field by using GS as the cumulus convection scheme and the AD formulation for the surface fluxes.

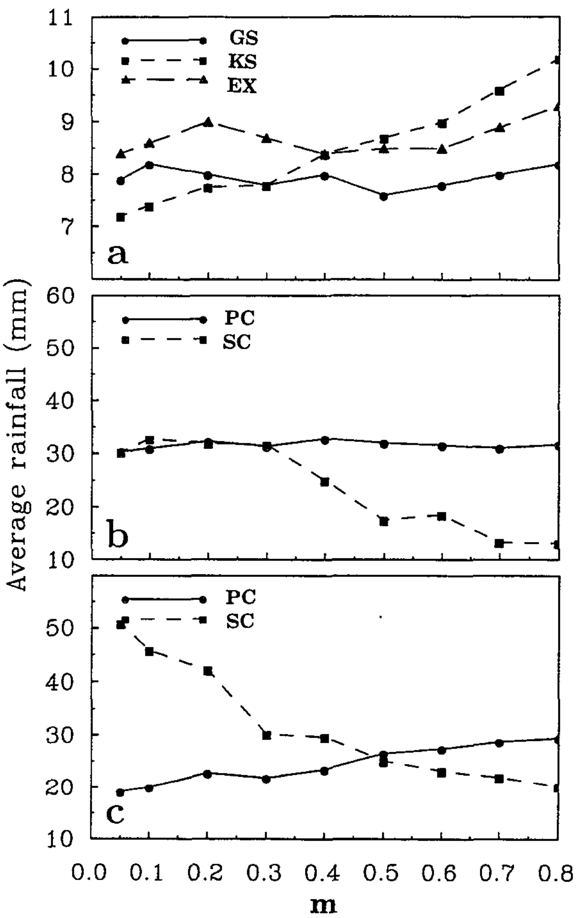

FIG. 6. Dependence of average 24-h accumulated rainfall (mm) simulated at 1200 UTC 9 July 1993 on m for different cumulus convection schemes: (a) averaged over the nested domain (EX denotes the fully explicit cloud scheme), (b) averaged over the PC and SC separately using GS, (c) same as (b) except KS was used.

The model produced a well-organized rainband oriented from east-northeast to west-southwest across Iowa, matching the position of the stationary front through the region. An elongated primary heavy rainfall center (denoted PC, primary center) was simulated in east-central Iowa. A secondary rainfall center (denoted SC, secondary center) with a peak value of 55 mm was simulated over eastern Nebraska. The shape and orientation of the simulated rainband were in close agreement with the observed pattern (Fig. 3b). The simulated peak rainfall (130 mm) matches well the observed peak (150 mm). However, its location was shifted about 100 km (three grid lengths) to the east of the observed. The simulated peak rainfall and location for the SC over eastern Nebraska were reasonably close to the observations.

The temporal distribution of simulated rainfall was also in close agreement with the observations. Two rainfall episodes, one late in the morning on 8 July and the other that night, accounted for rainfall at the PC. The second episode produced more rain than the first, but in the simulations the two episodes had similar magnitudes. Such overprediction during the early period (and underprediction during the late period) of a simulation has been reported as a deficiency of rainfall simulations (Cai et al. 1992). This deficiency seems to be attributed to the limitations of the cumulus parameterization schemes. Both observed and simulated rainfall for the SC developed late in the night of 8 July.

Simulations using a wide range of m values provide a systematic evaluation of soil-moisture impacts on rainfall for different cumulus convection schemes. Figure 4 presents the simulated 24-h accumulated rainfall fields for various values of m prescribed uniformly over the domain with GS as the cumulus convection scheme and the AD formulation for the surface flux scheme. The most notable feature is that as m increased, the PC intensified whereas the SC diminished. The rainfall at PC was more sensitive to changes in m when m was small (≤ 0.2), but as m increased, the rainfall fields remained essentially the same. This insensitivity for relatively high values of m is in agreement with one-dimensional studies (e.g., Kondo et al. 1990; Mahfouf and Noilhan 1991) implying that ET is most sensitive to soil moisture within a narrow range of low m values. The positive soil-moisture impact on rainfall for the PC is in agreement with the earlier findings of Pan et al. (1995). However, the SC over eastern Nebraska, which was not resolved in that study probably because of the coarse model and data resolutions, had an inverse dependence on m. For relatively dry surfaces ($m = 0.1$), the rainfall amount and areal coverage over the SC were only slightly less than those over the PC. However, when m increased, the rainfall area around the SC shrank and the magnitude decreased gradually, vanishing at $m = 0.6$. Also, as the rainfall associated with the SC diminished, the SC shifted toward the southwest, possibly as a result of the intensification of the nocturnal LLJ (see discussion later).

Figure 5 depicts the 24-h accumulated rainfall fields for different m values when KS was used instead of GS for cumulus convection. The general patterns of rainfall fields, including a northeast–southwest-oriented rainband with two centers, were similar to those produced by using GS. The major difference between the two schemes, however, is that GS produced spatially concentrated rainfall, whereas KS produced a widespread rainfall (for PC). The maximum rainfall at the PC, in the range of all m values, was only 55 mm, in contrast to 159 mm for the GS simulation. The impacts of soil moisture on rainfall were also similar to those for GS; the PC strengthened and the SC weakened with increasing m. However, the dependence of rainfall on m was more pronounced in KS. The PC noticeably strengthened over a large area with increasing m. This is in contrast to the GS results where an increase in peak rainfall, but little increase in areal coverage, was sim-

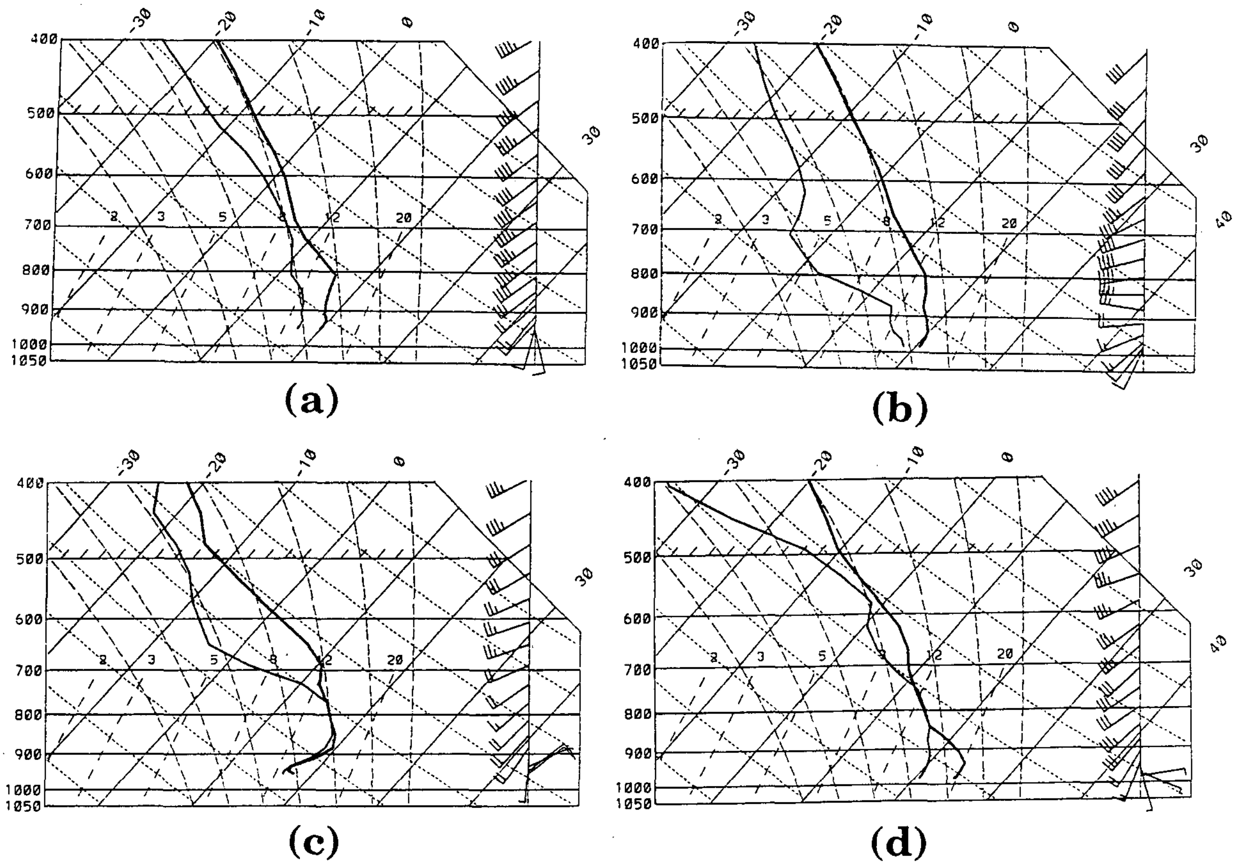

FIG. 7. Soundings at the PC (42.0°N, 93.0°W) and SC (42.3°N, 98.0°W): (a) and (b) for the SC and PC observed at 1200 UTC 8 July 1993, respectively, (c) and (d) for the SC and PC simulated at 0600 UTC 9 July 1993 with $m = 0.2$, respectively.

ulated. The PC barely existed in Iowa for $m \leqslant 0.2$. On the other hand, the SC strengthened considerably for the dry surface.

2) Rainfall averages

The simulated average rainfall over the nested domain (total rainfall over the entire nested domain divided by total grid points) was computed for various m values. In the GS simulation, the gridpoint average rainfall slightly decreased with m for low values of m (except for $m = 0.1$) but increased slightly for high values of m (Fig. 6a). This nonlinear behavior was due to the opposite responses of the two rainfall centers to m (see later). The gridpoint average rainfall over the domain was almost constant (~8 mm). On the other hand, average rainfall for the KS simulations increased linearly with m from about 7 to 10 mm. For $m \geqslant 0.3$, KS produced greater total rainfall than GS although the peak value at the PC for the KS simulation was less than one half of that from the GS results.

Figures 6b and 6c depict the average rainfall changes with m for the two rainfall centers (PC and SC) separately for GS and KS, respectively. (Rainfall was averaged over only those areas where the 24 h amounts exceeded 10 mm.) In GS, the average rainfall of the PC was about 30 mm for the entire range of m values. Rainfall at the SC was also about 30 mm for values of m less than 0.3, but for larger values of m rainfall diminished rapidly. For KS, the average rainfall for the PC linearly increased from 20 to 30 mm whereas average rainfall for the SC decreased from 50 to 20 mm over the same range of m values.

3) The elaboration on different responses of Kuo and Grell schemes to soil moisture

The reasons for the cumulus-scheme-sensitive response of rainfall to soil moisture are discussed in this subsection. However, it should be realized that isolating the processes responsible for the differences is difficult because of the highly nonlinear interactions involved.

One basis for the different responses relates to each scheme's closure assumptions. Convective adjustment of the environment in KS is constrained by a conservation of water substance requirement: convection is assumed to consume moisture at the rate (more accu-

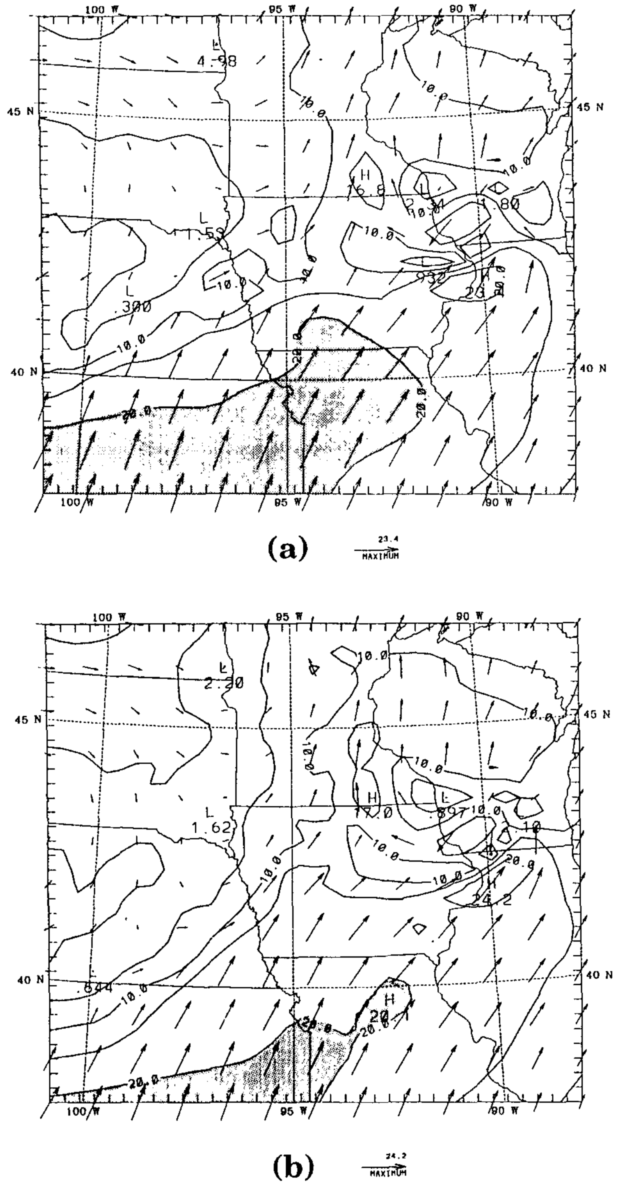

FIG. 8. Simulated wind speed (m s^{-1}) and direction at the level σ = 0.91 (about 700 m AGL) at 0600 UTC 9 July 1993 for (a) m = 0.2 and (b) m = 0.6. Areas with wind speeds above 20 m s^{-1} are stippled.

rately, at some fractional rate) at which it is supplied by the large-scale environment (Raymond and Emanuel 1993). Moisture convergence and ET contribute to this environmental supply, so for KS, moist processes directly influence convection, its adjustment of the environment, and the resultant precipitation. Convective adjustment in GS, on the other hand, is constrained by a conservation of available moist static energy: convection is assumed to consume the available moist static energy (i.e., available moist enthalpy) at the rate at which the large-scale environment supplies it. Since variations in soil moisture do not alter significantly the

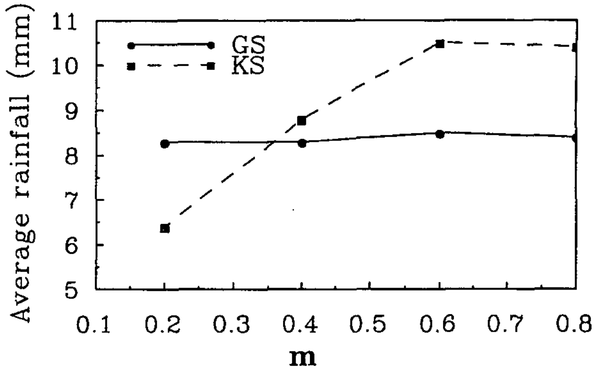

FIG. 9. Simulated domain-averaged 24-h accumulated rainfall (mm) at 1200 UTC 9 July 1993 using the SiB as surface flux for different cumulus parameterization schemes and m values.

surface moist enthalpy flux into the atmosphere (although it controls the partitioning between sensible and latent heat fluxes), the response of convection, and thus the convective rainfall, to soil moisture should be weaker for GS.

The further potential for differences in the response to soil moisture due to differences in the closure assumptions is illustrated by how each scheme treats excess moisture due to convergence. For KS, a fraction of the excess moisture will be instantly [i.e., within a single time step (90 s)] rained out, while the remainder will be vertically redistributed in the atmospheric column. For GS, excess moisture is not necessarily rained out instantly because precipitation is only influenced by such excessive moisture indirectly through the cloud work function (Arakawa and Schubert 1974). There-

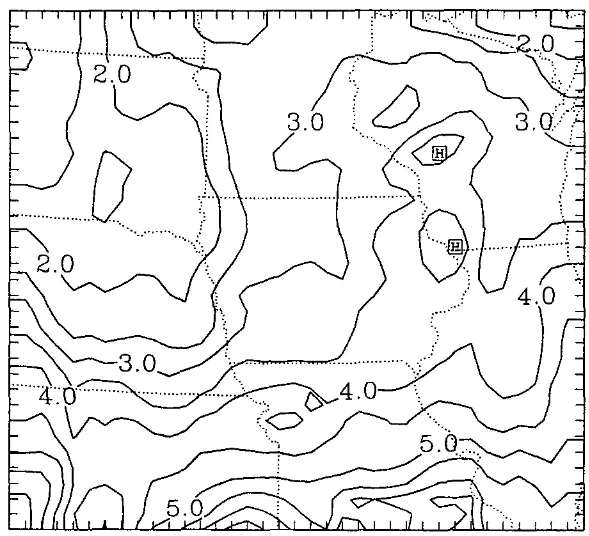

FIG. 10. Simulated 24-h ET (mm) distribution at 1200 UTC 9 July 1993 with estimated m field given in Fig. 1 using the AD formulation and GS scheme.

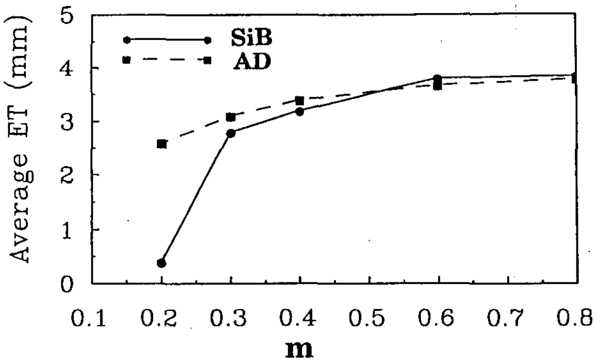

FIG. 11. Domain-averaged 24-h ET at 1200 UTC 9 July 1993 for different m values using the SiB and AD schemes.

fore, convective rainfall in KS responds more directly and more rapidly to the excess moisture than GS.

Another possible explanation for the greater sensitivity of rainfall to m in KS relates to differences in how each scheme partitions total rainfall between subgrid (convective) and resolvable scale (stratiform) forms. For this case in KS, precipitation was mostly subgrid scale, whereas in GS it was mostly resolvable. It is generally accepted that convective precipitation is more strongly dependent on surface forcing (e.g., Benjamin and Carlson 1986), and thus it is expected that KS would be more sensitive to surface forcing for this case.

To explore the effects of rainfall partitioning between subgrid and resolvable scales, further simulations were conducted in which precipitation was treated as being totally resolvable. The use of only a fully explicit cloud scheme (EX) can be partly justified considering that the rainfall system was quite large in this case, although the model horizontal resolution of 30 km is relatively coarse (Rosenthal 1978). The results from these fully explicit cloud simulations were very similar to those produced by GS except that they produced about 0.8 mm more rainfall averaged over the domain (Fig. 6a). This result supports the argument that the reduced sensitivity of rainfall to m in GS is because it partitions more precipitation into the resolvable form.

Given the reduced sensitivity of simulated rainfall to m when the fully explicit cloud scheme was used and the more rapid and direct response of KS to excess moisture, it is suggested that rainfall exhibits an *exaggerated* sensitivity to soil moisture for the KS simulations. This exaggerated sensitivity is responsible for the major differences in the soil-moisture impacts on the simulated rainfall between the cumulus parameterizations.

4) EFFECTS OF LOCAL CONDITIONS

Two possible causes are offered to explain the inverse response of rainfall to m at the SC. The first is related to the difference in the temperature and humidity stratification between the PC and SC. Figure 7 shows the soundings at these two rainfall centers at the initial time (1200 UTC 8 July, Figs. 7a,b) and 18 h into the integration (0600 UTC 9 July, Figs. 7c,d), when rainfall at the SC was near peak intensity. (Note the KS and AD schemes were used and m was set to 0.2 in this simulation.) The main difference was that the atmosphere was more moist and stable at the SC where a deep stable layer extended up to 800 hPa. Thus, the relatively humid and thermally stable atmosphere at the SC had enough moisture to sustain convection, but it lacked sufficient sensible heating to break the inversion. As m increased, the increased ET reduced the sensible heat flux, which, in turn, suppressed the development of the convective boundary layer (CBL). More sensible heating, not moistening, is needed to increase rainfall (Otterman et al. 1990; Fuelberg et al. 1991; Sanders and Blanchard 1993). At the PC, on the other hand, the atmosphere was dry and nearly neutral, and the increased ET fueled convection and enhanced rainfall as convection developed during the day.

The contrasting behaviors at PC and SC suggests that an increase in ET enhances rainfall if a dry and well-developed CBL has already been established, but it decreases rainfall if ET suppresses the development of surface-based convection. It is worth noting that increased soil moisture also can suppress the development of fair weather boundary layer clouds (Ek and Mahrt 1994; Segal et al. 1995).

The second cause for the decreased rainfall with increasing m at the SC may be linked to the nocturnal LLJ. Nocturnal southwesterly flow at about 700 m AGL, around which the LLJ typically reaches its peak intensity (Mitchell et al. 1995), was stronger to the south of the SC for $m = 0.2$ than for $m = 0.6$ (Fig. 8). Thus for $m = 0.2$, there was increased advection of moisture, from the south, into the region. The simulated intensification of the LLJ with decreasing m is in agreement with the results of McCorcle (1988) and Turner (1993). The intensification is caused by the increased daytime heating of the CBL for the drier surfaces. As a result, the nocturnal wind strengthened and consequently the southerly flow intensified.

c. Sensitivity of simulated rainfall to surface flux schemes

The aerodynamic formulation, AD, and the SiB scheme were used to evaluate the sensitivity of soil moisture impacts to surface flux schemes. Overall, rainfall fields simulated using the SiB scheme resembled those produced using the AD scheme for both GS and KS. The change in domain average rainfall with m, when SiB was incorporated into the model as the flux scheme, is depicted in Fig. 9. For $m \leqslant 0.6$, average rainfall simulated with KS increased linearly with m,

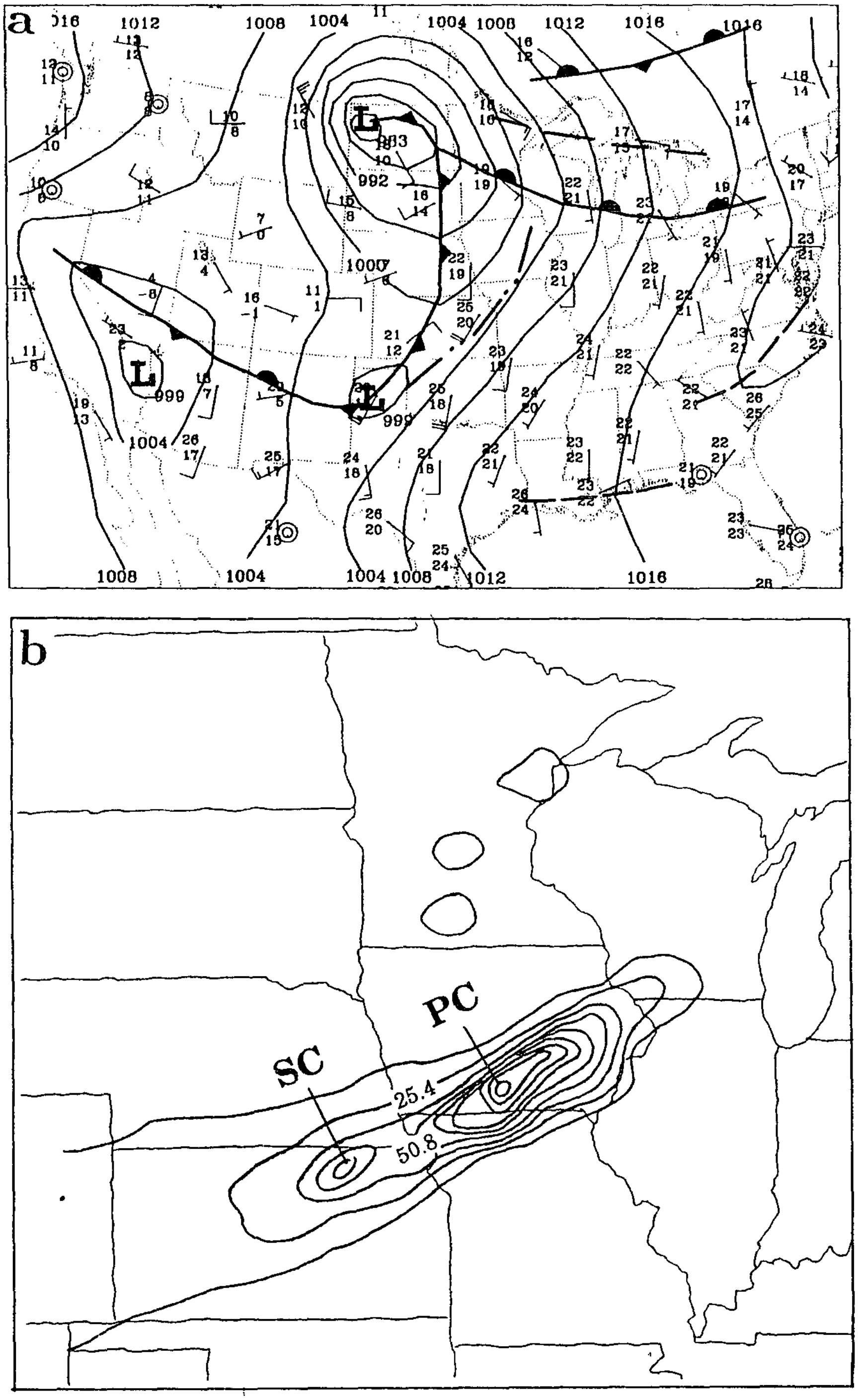

FIG. 12. (a) Surface synoptic chart at 1200 UTC 4 July 1993. The dashed lines are troughs and the dash–dotted line is the squall line. (b) Same as Fig. 3b except at 1200 UTC 5 July 1993.

whereas that simulated with GS remained constant. This is similar to the results obtained with the AD scheme (see Fig. 6a), and it suggests that surface flux schemes had only a secondary influence on soil-moisture effects. One difference, however, is that for simulations in which SiB was used rainfall did not change when $m \geqslant 0.6$ (even for KS). This could be caused by differences in the spatial distribution of ET (discussed in the next section) as computed by the two flux schemes.

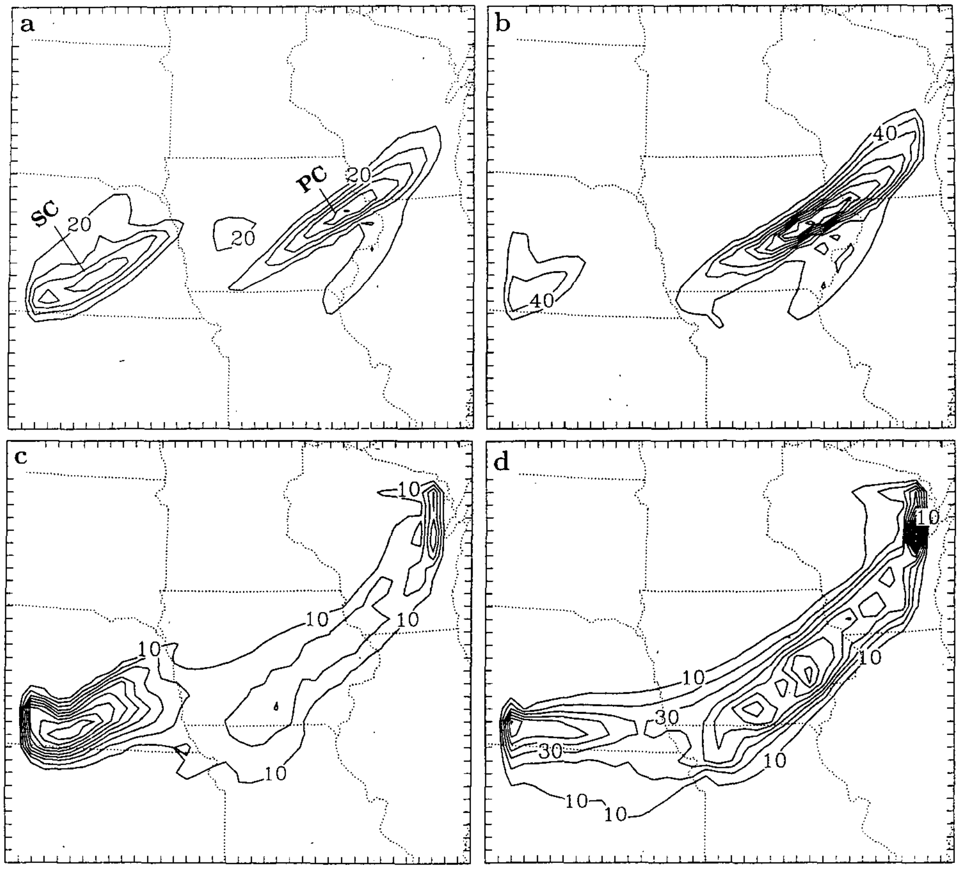

FIG. 13. Simulated 24-h accumulated rainfall (mm) at 1200 UTC 5 July 1993 by GS and KS for different m values: (a) and (b) GS for $m = 0.2$ and $m = 0.6$, respectively, (c) and (d) KS for $m = 0.2$ and $m = 0.6$, respectively; PC and SC denote the primary and secondary rainfall centers, respectively.

A modified AD formulation, which has been shown to reduce the overestimation of surface moisture flux (Pan et al. 1994), also was compared with the conventional (MM5) formulation to estimate possible effects of the overestimation on rainfall. Results indicated that the use of the alternative formulation had little influence on soil moisture effects.

4. Sensitivity of evapotranspiration to different surface flux schemes—9 July 1993 case

Figure 10 presents the daily ET field calculated using the AD formulation with the estimated m field as shown in Fig. 1. The dependence of ET on m is modulated by clouds, surface temperature, and surface layer humidity. On this day, a minimum ET of about 2 mm was simulated over eastern Nebraska and South Dakota due to lower m values (see Fig. 1), and ET values of 4.0–6.5 mm were simulated in the southern part of the domain. The predicted daily ET over Iowa, where the heavy rainfall event occurred, of about 3 mm was lower than the climatological value of 4 mm (Shaw 1981) because of the above average cloudiness.

There are no directly observed values of ET to verify the simulations. However, by multiplying the pan evaporation by a conversion factor of 0.77 (Shaw 1981), crude estimates of actual ET can be obtained. Pan evaporation measurements available from three locations in Iowa on 9 July gave an average ET of approximately 3.5 mm. This indicates that the simple AD formulation produced reasonable ET values over Iowa on this particular day.

Figure 11 shows the average ET in the entire nested domain over the 24-h period for various m values. Uniform values of m were specified over the entire domain for this set of experiments. Both SiB and AD schemes showed that ET increased rapidly with m when m was small and then leveled off as m increased further. The two schemes showed no significant difference when $m \geqslant 0.3$, but ET was lower for the SiB scheme when $m \leqslant 0.2$. This was in part because soil moisture approached the wilting point (i.e., a condition where evapotranspiration nearly

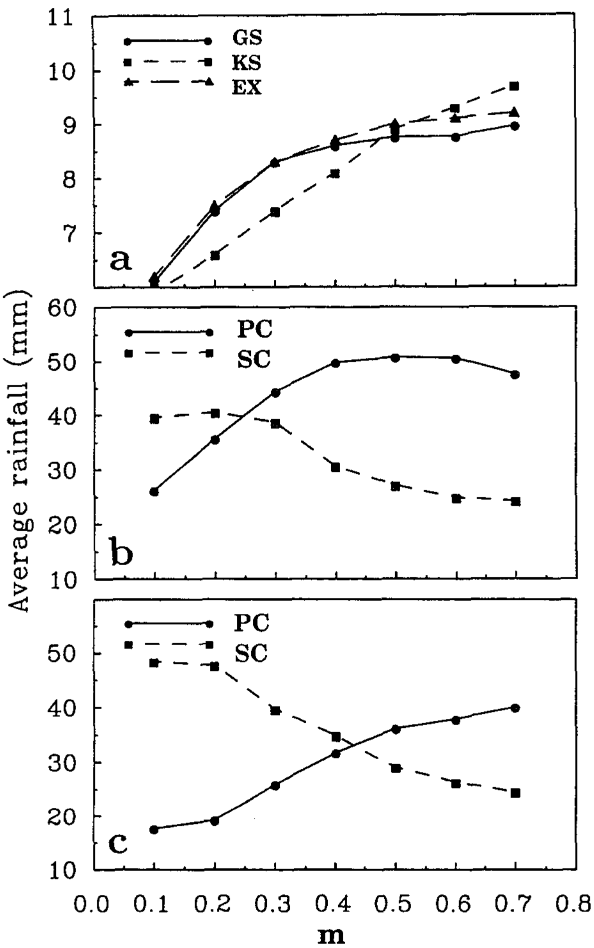

FIG. 14. Same as Fig. 6 except at 1200 UTC 5 July 1993.

stops) in the SiB scheme. The AD formulation in this study has no constraints to account for the wilting point. Another reason for the reduced sensitivity of ET to m when $m \leq 0.2$ in this AD formulation is that the surface specific humidity is assumed to be proportional to soil temperature. The AD formulation tends to overestimate ET for high soil temperatures.

The spatial distributions of ET were different between the two schemes even though their domain-averaged ETs were very close. The AD scheme produced high values of ET over the southwest corner and low values over the northern part of the domain because of the two previously described effects (i.e., wilting point and temperature).

The differences in ET in the range of $m = 0.2$–0.8 were only 1 mm over the PC area (not shown). The maximum differences in simulated rainfall over the PC, however, changed by 8 mm (Fig. 6c). The rainfall enhancement was, therefore, about eight times as large as the enhancement of local ET. In other words, the increased rainfall at one grid cell at the rainfall center was equal to the total increased ET contributed from eight neighboring grid cells. The total rainfall over the whole domain (nested only) for KS increased by 4.0 mm (Fig. 9), whereas the total ET increased only by 1.2 mm (Fig. 11) as m increased from 0.2 to 0.8. This result suggests that over a domain of 10^6 km^2, the rainfall increased three to four times more than the ET had.

5. Further simulations—5 July and 17 June 1993 cases

The base simulations for 9 July suggested that the soil moisture impacts on rainfall were more strongly dependent on cumulus schemes than on surface moisture flux schemes. Furthermore, simulated rainfall in KS was noticeably more sensitive to m than in GS. To further test these findings, two additional events were examined.

The upper-air synoptic pattern for the 5 July case was very similar to that for 9 July. At 1200 UTC 4 July a deep surface low pressure area was centered in north-western South Dakota (Fig. 12a). A squall line almost parallel with the cold front extended from southeast South Dakota to the Oklahoma panhandle. A narrow band of heavy rain with two maxima (Fig. 12b) was produced along the squall line. In this case, both GS and KS basically produced the rainband structure, especially for $m = 0.6$ (Fig. 13); however, KS was more sensitive to m. The average rainfall at the PC increased by about a factor of 3 as m increased from 0.2 to 0.6 for KS. On the other hand, for GS, rainfall increased by only 30%, although the absolute increase was larger than in KS.

Rainfall averaged over the nested domain increased linearly with m for KS, whereas for GS (and EX) it increased rapidly for $m \leq 0.3$ and leveled off for larger m (Fig. 14a). Rainfall at the PC for GS increased with m first and then leveled off (Fig. 14b), whereas for KS rainfall at the PC increased almost linearly with m (Fig. 14c). The slight drop in rainfall at PC for GS at large m values was due to the expansion of rain areas without total rainfall increase. The overall sensitivity of rainfall to m for the two rainfall centers (PC and SC) for both cumulus convection schemes was very similar to that of the 9 July case.

The second event, 17 June, was selected to further evaluate model sensitivities. The upper-air flow pattern around 17 June was characterized by a strong zonal flow that provided a "duct" for an intense cyclone propagating into the Midwest from the western Pacific (Gerald and Janowiak 1995). The surface low pressure associated with the cyclone was located in central Nebraska at 1200 UTC 16 June (Fig. 15a). The system produced heavy rainfall (Fig. 15b) as it moved eastward during the next 24 h. The rainfall fields simulated using GS and KS

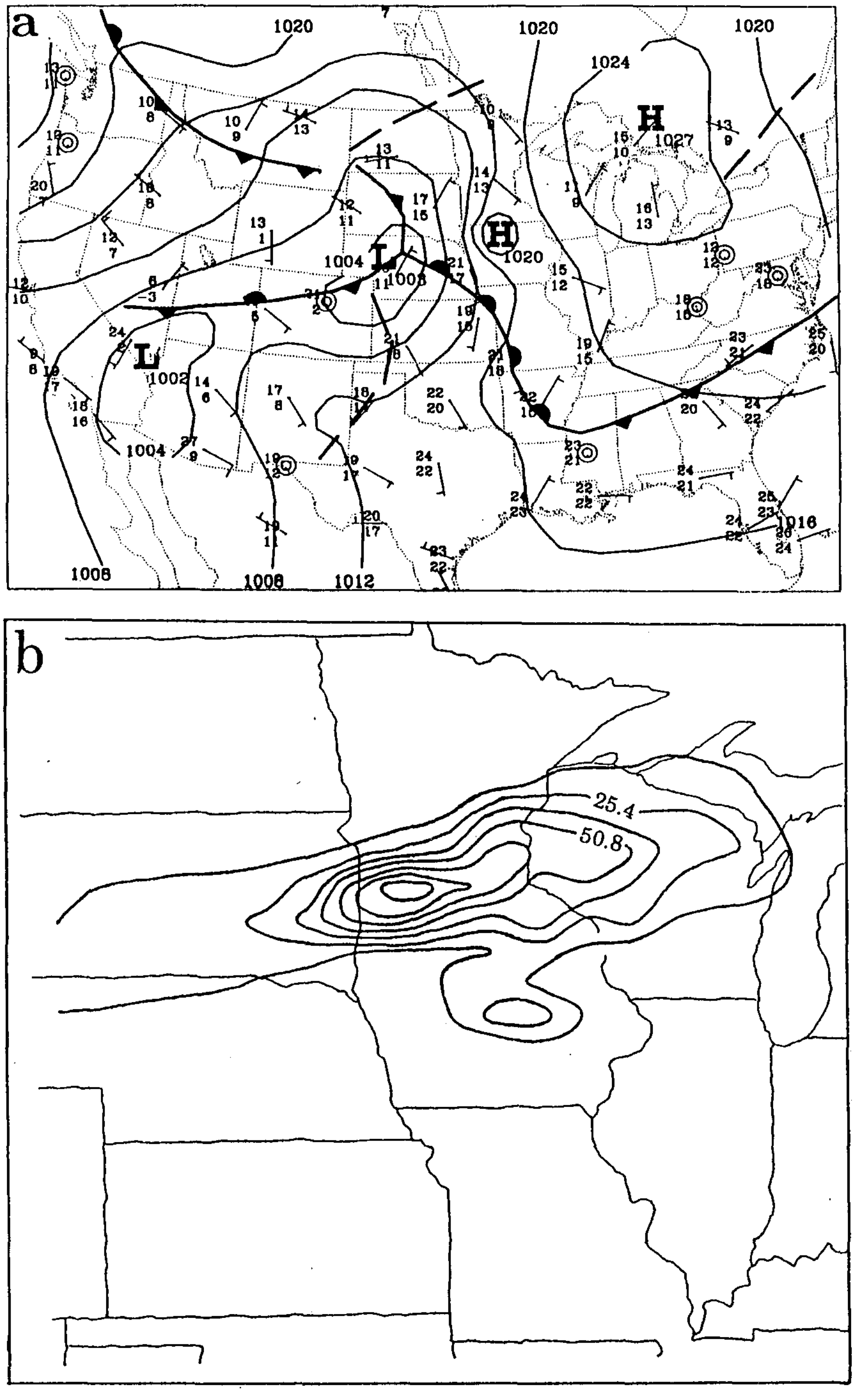

FIG. 15. (a) Surface synoptic chart at 1200 UTC 16 June 1993. The dashed lines are troughs. (b) Same as Fig. 3b except at 1200 UTC 17 June 1993.

for m = 0.2 and 0.6 are presented in Fig. 16. The rainband for GS was affected slightly by the change in m. In contrast, the rainband for KS was hardly recognizable when m = 0.2, whereas a well-organized rainband was produced with m = 0.6. The domain-averaged rainfall in this event was noticeably more sensitive to m for KS than for either GS or EX (Fig. 17).

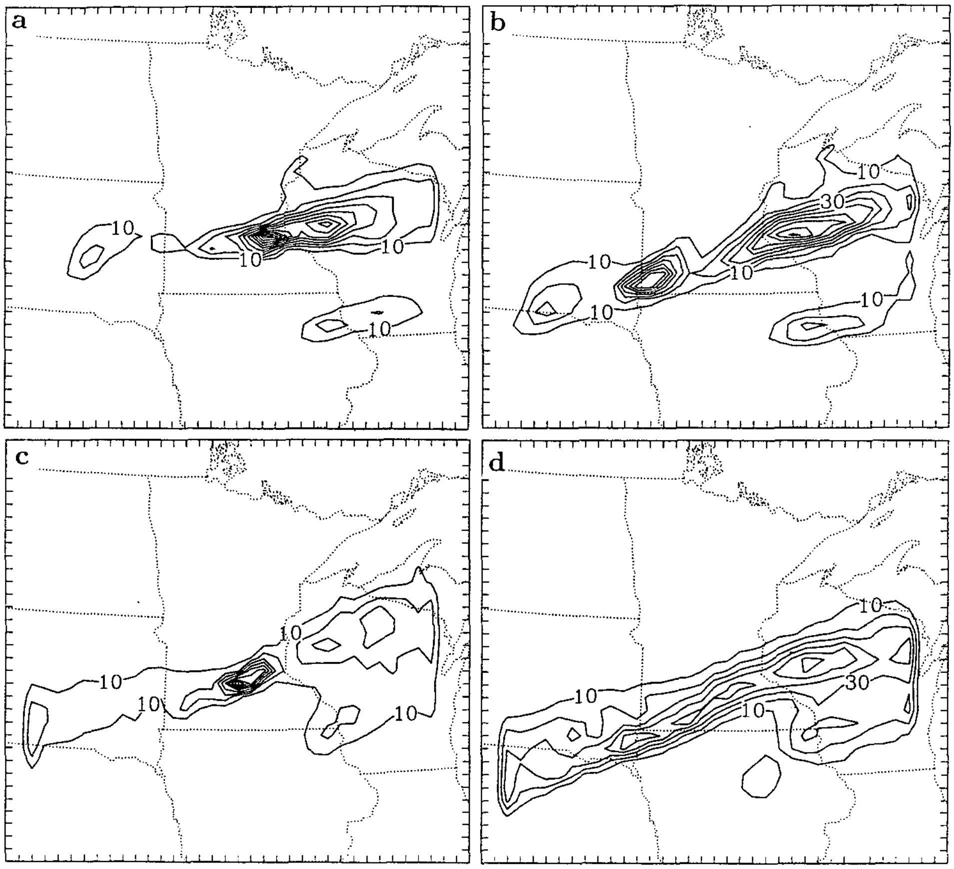

FIG. 16. Simulated 24-h accumulated rainfall (mm) at 1200 UTC 17 June 1993 by GS and KS for different m values: (a) and (b) GS for $m = 0.2$ and $m = 0.6$, respectively, (c) and (d) KS for $m = 0.2$ and $m = 0.6$, respectively.

6. Summary and conclusions

The present study evaluates influences of cumulus parameterization and surface flux schemes on summer rainfall response to transient soil moisture changes over the central United States. Simulations of 24-h duration allowed for a large number of realizations (various m values) for each event. The cumulus convection schemes used were the Kuo scheme, Grell scheme, and a fully explicit cloud physics scheme. Sensitivities to surface flux schemes was tested by comparing the results from the aerodynamic formulation to those with the Simple Biosphere Model.

Increased soil moisture was found to enhance domain total rainfall but to enhance or suppress local rainfall, depending on local thermal and moisture conditions of the atmosphere. For two events (5 and 9 July 1993), each having two separate rainfall centers, the primary centers located in eastern Iowa strengthened as m increased, whereas the secondary rainfall centers located in eastern Nebraska weakened. It was found that increased evapotranspiration (ET) from a wet surface reduced local rainfall when the lower atmosphere was humid and lacked sufficient thermal forcing to initiate deep convection. On the other hand, when the lower atmosphere was thermally unstable but relatively dry,

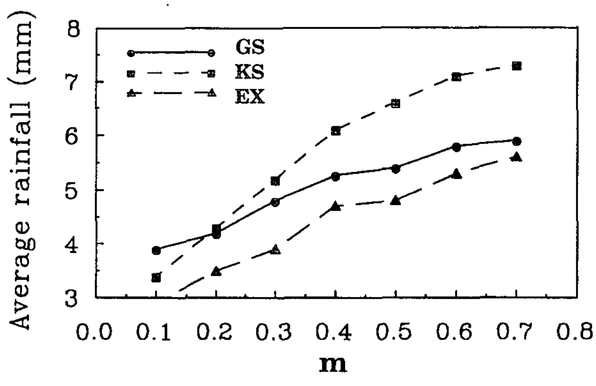

FIG. 17. Dependence of simulated domain-averaged 24-h accumulated rainfall (mm) on m at 1200 UTC 17 June 1993 for different cumulus convection schemes (EX denotes the fully explicit cloud scheme).

an increase in soil moisture increased local rainfall. It may therefore be speculated that during the early stage of convective development when the convective boundary layer is still shallow, increased ET may suppress convection and thus precipitation. On the other hand, after a deep unstable layer has been already established, increased ET may fuel convection by adding more moisture and thus enhance precipitation. In other words, sensible heating is more important in initiating convection and ET is more important in amplifying convection (precipitation). More studies are needed to confirm this conjecture.

The nocturnal low-level jet, a major mechanism influencing heavy rainfall in this region, is an additional complicating factor in determining precise soil-moisture impacts on rainfall for the cases simulated in this study. The nocturnal low-level jet weakened with increasing m. This reduced the transport of moisture into the region, and contributed to the decrease in rainfall at the secondary center for high m values.

Soil moisture impacts were noticeably stronger for the Kuo scheme, which produced lighter peak rainfall, than those for the Grell scheme, which simulated heavier peak rainfall. The impacts were roughly linear with soil moisture availability m for the Kuo scheme, whereas for the Grell scheme they were nonlinear and varied among cases. The difference in rainfall response to m between the two schemes was attributed, in part, to how each scheme partitioned total rainfall between the subgrid and resolvable scales. The different responses could also be attributed to differences in the parameterization formulations, such as closure assumptions. For example, a fraction of any ET contributing to excess moisture is directly rained out in the Kuo scheme because moisture is assumed to be consumed by convection at the rate at which it is supplied. However, this is not the case for the Grell scheme because the available moist static energy, not moisture, is conserved.

The fully explicit cloud physics simulations (without a cumulus parameterization scheme) showed a significantly weaker response of rainfall to m than the Kuo scheme, which produced rainfall mostly in subgrid convective form. This suggests that a cumulus convection parameterization scheme like the Kuo scheme can exaggerate the soil moisture impacts on rainfall. Since most previous soil moisture impact simulations used only one cumulus parameterization scheme (the Kuo scheme has been most frequently used), the impacts could have been exaggerated.

The simple but widely used aerodynamic formulation and the more physically sound SiB scheme produced similar rainfall responses to m, suggesting that details of the surface flux scheme have only a secondary effect on transient soil-moisture impacts compared with those of cumulus convection schemes.

Acknowledgments. This is a journal paper J-16466 of the Iowa Agriculture and Home Economics Experiment Station, Ames, Iowa (Project 3153). Partial support was provided by NASA Grant NAG 5-2491 and NSF Grant ATM-9319455. Data and computer support used in this study were partly provided by the Science Computing Division of the National Center for Atmospheric Research sponsored by the NSF. The authors wish to thank Dr. Sue Chen for valuable consulting on the MM5 code and Dr. James Collatz for providing the SiB2 code and some input parameters. We would like also to thank Drs. John Brown and Ray Arritt for helpful discussions and Richard Carlson for providing climatological data.

APPENDIX

Outline of Evapotranspiration Computation in SiB Model

The SiB2 model (Sellers et al. 1992; Collatz et al. 1990) is the improved version of the original SiB model. It uses a semianalytical form of integrated canopy resistance that is proportional to photosynthesis rate. The stomatal conductivity is given by

$$g_s = M \frac{A_n}{C_s} h_s P_s + b, \qquad \text{(A1)}$$

where M and b are constants, C_s is the CO_2 concentration, h_s is an adjustment factor for relative humidity of air at the surface, P_s is the surface pressure, and A_n is the photosynthesis rate.

The canopy is divided into an upper story, which represents tall trees and plants, and lower story representing subcanopy grass. The soil is represented by a thin top layer, middle root layer, and deep layer below. The moisture flux from the surface consists of evapotranspiration from canopy (E_c) and ground cover (E_g) (Sellers et al. 1986):

$$E_c = [e(T_c) - e_a] \frac{\rho c_p}{\gamma L_\nu} \left(\frac{W_c}{\overline{r_b}} + \frac{1 - W_c}{\overline{r_b} + \overline{r_c}} \right), \qquad \text{(A2a)}$$

$$E_g = [e(T_g) - e_a] \frac{\rho_a c_p}{\gamma L_\nu} \left[\sigma_g \left(\frac{W_g}{r_d} + \frac{1 - W_g}{r_d + r_g} \right) + \frac{(1 - \sigma_g) h_s}{r_{\text{surf}} + r_d} \right], \qquad \text{(A2b)}$$

where T_c is the canopy temperature; $e(T_c)$ and $e(T_g)$ are the saturation vapor pressure at T_c and T_g, respectively; e_a is the ambient water vapor pressure; W_c and W_g are the wetted fractions of canopy and ground cover, respectively (W_c and W_g are assumed to zero in this study); σ_g is the ground fraction covered by vegetation (herein assumed to be 0.79); $\overline{r_b}$ and $\overline{r_c}$ are the bulk boundary layer and canopy resistance, respectively; r_d is the aerodynamic resistance between ground and air at canopy source height; r_g is the bulk stomatal resistance of ground cover; r_{surf} is the bare soil surface resistance, γ is the psychrometric constant, L_ν is the latent heat of condensation, and c_p is the specific heat at constant pressure.

In SiB2, the predicted soil water content η is defined as the ratio of actual to saturation volumetric water content of soil. The soil moisture availability m also can be approximated as this ratio (Deardorff 1972; Manabe 1969). Thus, in our application, it is assumed that η defined in SiB2 and m used in the aerodynamic formula have equivalent effects on evapotranspiration. Considering the ambiguity associated with the definition of m (Davies and Allen 1973), this rough equivalence is justified. To be consistent with uniformly specified soil moisture, the vegetation over the entire domain was assumed as a mixture of crop and grass fields grown on loam soil. This assumption is reasonable since our domain of interest is over the United States midwest, an intensively cultivated region.

REFERENCES

Anthes, R. A., 1977: A cumulus parameterization scheme utilizing a one-dimensional cloud model. *Mon. Wea. Rev.*, **105**, 270–286.

Arakawa, A., and W. H. Schubert, 1974: Interaction of a cumulus cloud ensemble with the large-scale environment, Part I. *J. Atmos. Sci.*, **31**, 674–701.

Atlas, R., N. Wolfson, and J. Terry, 1993: The effect of SST and soil moisture anomalies on GLA model simulations of the 1988 U.S. summer drought. *J. Climate*, **6**, 2034–2048.

Augustine, J. A., and F. Caracena, 1994: Lower-tropospheric precursors to nocturnal MCS development over the central United States. *Wea. Forecasting*, **9**, 116–135.

Benjamin, S. G., and T. N. Carlson, 1986: Some effects of surface heating and topography on the regional severe storm environment. Part I: Three-dimensional simulations. *Mon. Wea. Rev.*, **114**, 307–329.

Blackadar, A. K., 1976: Modeling the nocturnal boundary layer. Preprints, *Third Symp. on Atmospheric Turbulence, Diffusion and Air Quality*, Raleigh, NC, Amer. Meteor. Soc., 46–49.

Cai, Z. Y., Z. S. Wang, and Z. T. Pan, 1992: A numerical study on forecasting the Henan extraordinary heavy rainfall event in August 1975. *Adv. Atmos. Sci.*, **9**, 53–62.

Chang, J.-T., and P. J. Wetzel, 1991: Effects of spatial variations of soil moisture and vegetation on the evolution of a prestorm environment: A numerical case study. *Mon. Wea. Rev.*, **119**, 1368–1390.

Collatz, J. G., J. A. Berry, G. D. Farquhar, and J. Pierce, 1990: The relationship between the rubisco reaction mechanism and models of photosynthesis. *Plant Cell Environ.*, **13**, 219–225.

Davies, J. A., and C. D. Allen, 1973: Equilibrium, potential and actual evaporation from cropped surfaces in southern Ontario. *J. Appl. Meteor.*, **12**, 649–657.

Deardorff, J. W., 1972: Parameterization of the planetary boundary layer for use in general circulation models. *Mon. Wea. Rev.*, **100**, 93–106.

Ek, M., and L. Mahrt, 1994: Daytime evolution of relative humidity at the boundary layer top. *Mon. Wea. Rev.*, **122**, 2709–2721.

Fritsch, J. M., and C. F. Chappell, 1980: Numerical prediction of convectively driven mesoscale pressure systems. Part I: Convective parameterization. *J. Atmos. Sci.*, **37**, 1722–1733.

Fuelberg, H. E., R. L. Schudalla, and A. R. Guillory, 1991: Analysis of sudden mesoscale drying at the surface. *Mon. Wea. Rev.*, **119**, 1391–1406.

Gerald, D. B., and J. E. Janowiak, 1995: Atmospheric circulation associated with the Midwest flood of 1993. *Bull. Amer. Meteor. Soc.*, **76**, 681–685.

Giorgi, F., 1991: Sensitivity of simulated summertime precipitation over the western United States to different physics parameterizations. *Mon. Wea. Rev.*, **119**, 2870–2888.

——, M. R. Marinucci, G. T. Bates, and G. De Canio, 1993: Development of a second-generation regional climate model (RegCM2). Part II: Convective processes and assimilation of boundary conditions. *Mon. Wea. Rev.*, **121**, 2814–2832.

Grell, G. A., 1993: Prognostic evaluation of assumptions used by cumulus parameterizations. *Mon. Wea. Rev.*, **121**, 764–787.

——, J. F. Dudhia, and D. Stauffer, 1993: A description of the fifth generation Penn State/NCAR Mesoscale Model (MM5). NCAR Tech. Note NCAR/TN-398+IA, 107 pp. [Available from NCAR, P.O. Box 3000, Boulder, CO 80307.]

Hsie, E.-T., and R. A. Anthes, 1984: Simulations of frontogenesis in a moist atmosphere using alternative parameterizations of condensation and precipitation. *J. Atmos. Sci.*, **41**, 2701–2716.

Kondo, J., N. Saigusa, and T. Sato, 1990: A parameterization of evaporation from bare soil surfaces. *J. Appl. Meteor.*, **29**, 385–389.

Kuo, S. L., 1974: Further studies of the parameterization of the effect of cumulus convection on large-scale flow. *J. Atmos. Sci.*, **31**, 1232–1240.

Kuo, Y.-H., and R. A. Anthes, 1984: Semiprognostic tests of Kuo-type cumulus parameterization schemes in an extratropical convective system. *Mon. Wea. Rev.*, **112**, 1498–1509.

Lakhtakia, M. N., and T. T. Warner, 1987: A real-data numerical study of the development of precipitation along the edge of an elevated mixing layer. *Mon. Wea. Rev.*, **115**, 156–168.

Lanicci, J. M., T. N. Carlson, and T. T. Warner, 1987: Sensitivity of the Great Plains severe-storm environment to soil-moisture distribution. *Mon. Wea. Rev.*, **115**, 2660–2673.

Mahfouf, J. F., and J. Noilhan, 1991: Comparative study of various formulations of evaporation from bare soil using in situ data. *J. Appl. Meteor.*, **30**, 1354–1365.

Manabe, S., 1969: Climate and the ocean circulation. I: The atmospheric circulation and the hydrology of the earth's surface. *Mon. Wea. Rev.*, **97**, 739–774.

McCorcle, M. D., 1988: Simulation of surface-moisture effects on the Great Plains low-level jet. *Mon. Wea. Rev.*, **116**, 1705–1720.

Mintz, Y., 1984: The sensitivity of numerically simulated climates to land-surface boundary conditions. *The Global Climate*, J. T. Houghton, Ed., Cambridge University Press, 79–105.

Mitchell, M. J., R. W. Arritt, and K. Labas, 1995: A climatology of warm season Great Plains low-level jet using wind profiler observations. *Wea. Forecasting*, **10**, 576–591.

Mo, K. C., J. N. Paegle, and J. Paegle, 1995: Physical mechanisms of the 1993 floods. *J. Atmos. Sci.*, **52**, 879–895.

Molinari, J., and M. Dudek, 1992: Parameterization of convective precipitation in mesoscale numerical models: A critical review. *Mon. Wea. Rev.*, **120**, 326–344.

Otterman, J., A. Manes, S. Rubin, P. Alpert, and D. O'C. Starr, 1990: An increase of early rains in southern Israel following land-use change? *Bound.-Layer Meteor.*, **53**, 333–351.

Pan, Z., S. G. Benjamin, J. M. Brown, and T. Smirnova, 1994: Comparative experiments with MAPS on different types of parameterization schemes for surface moisture flux and boundary-layer processes. *Mon. Wea. Rev.*, **122**, 449–470.

——, M. Segal, R. Turner, and E. Takle, 1995: Model simulation of impacts of transient surface wetness on summer rainfall in the United States midwest during drought and flood years. *Mon. Wea. Rev.*, **123**, 1575–1581.

Raymond, D. J., and K. A. Emanuel, 1993: The Kuo cumulus parameterization. *The Representation of Cumulus Convection in Numerical Models, Meteor. Monogr.*, No. 46, Amer. Meteor. Soc., 145–147.

Richards, F., and M. D. Hudlow, 1994: Hydrometeorological conditions during the Great Flood of 1993. Preprints, *Symp. on the Great Flood of 1993*, Nashville, TN, Amer. Meteor. Soc., 65 pp.

Rodenhuis, D. R., D. Miskus, G. D. Bell, and K. C. Mo, 1994: Meteorological flood-origin, description, and causes of the Great Flood of 1993. Preprints, *Symp. on the Great Flood of 1993*, Nashville, TN, Amer. Meteor. Soc., 65 pp.

Rosenthal, S. L., 1978: Numerical simulation of tropical cyclone development with latent heat release by resolvable scales. I: Model description and preliminary results. *J. Atmos. Sci.,* **35,** 258–271.

Rowntree, P. R., and J. R. Bolton, 1983: Simulation of the atmospheric response to soil moisture anomalies over Europe. *Quart. J. Roy. Meteor. Soc.,* **109,** 501–526.

Sanders, F., and D. O. Blanchard, 1993: The origin of a severe thunderstorm in Kansas on 10 May 1985. *Mon. Wea. Rev.,* **121,** 133–149.

Sato, N., P. J. Sellers, D. A. Randall, E. K. Schneider, J. Shukla, T. L. Kinter III, Y.-T. Hou, and E. Albertazzi, 1989: Effects of implementing the simple biosphere model in a general circulation model. *J. Atmos. Sci.,* **46,** 2757–2782.

Segal, M., R. W. Arritt, C. Clark, R. Rabin, and J. M. Brown, 1995: Scaling evaluation of the effect of surface characteristics on potential for deep convection over uniform terrain. *Mon. Wea. Rev.,* **123,** 383–400.

Sellers, P. J., Y. Mintz, Y. C. Sud, and A. Dalcher, 1986: The design of a simple biosphere model (SiB) for use within general circulation models. *J. Atmos. Sci.,* **43,** 505–531.

——, J. A. Berry, G. J. Collatz, C. B. Field, and F. G. Hall, 1992: Canopy reflectance, photosynthesis, and transpiration. III. A reanalysis using improved leaf models and a new canopy integration scheme. *Remote Sens. Environ.,* **42,** 187–216.

Shaw, R. H., 1981: Evaporation climatology of Iowa. Special Rep. 88, Agriculture and Home Economics Experiment Station, Iowa State University, 20 pp. [Available from Iowa State University, Ames, IA 50011.]

Turner, R. W., 1993: The impact of climate change on the Great Plains low-level jet. Ph.D. dissertation, Iowa State University, 116 pp.

Wang, W., and N. Seaman, 1994: A comparative study of cumulus convection schemes in MM5. Preprints, *Tenth Conference on Numerical Weather Prediction,* Portland, OR, Amer. Meteor. Soc. 99–101.

WGNE, 1992: Report of the seventh session of the CAS/JSC working group on numerical experimentation. WMO.TD-477, World Meteorological Organization, Geneva, Switzerland, 50 pp.

Yan, H., and R. A. Anthes, 1988: The effect of variations in surface moisture on mesoscale circulations. *Mon. Wea. Rev.,* **116,** 192–208.

Zhang, D.-L., and R. A. Anthes, 1982: A high resolution model of the planetary boundary layer-sensitivity tests and comparisons with SESAME-79 data. *J. Appl. Meteor.,* **21,** 1594–1609.

——, E.-Y. Hsie, and M. W. Moncrieff, 1988: A comparison of explicit and implicit prediction of convective and stratiform precipitating weather systems with a meso-scale numerical model. *Quart. J. Roy. Meteor. Soc.,* **114,** 31–60.